

1 Late Cretaceous winter sea ice in Antarctica?

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8 ABSTRACT

9 The Late Cretaceous is generally considered to have been a time of greenhouse climate,
10 with no direct geological evidence for glaciation. We present, indirect evidence from the
11 Maastrichtian sedimentary record for significant, rapid sea-level changes suggests that
12 ice sheets were growing and decaying on Antarctica at that time. Evidence for possible
13 seasonal sea ice during the Maastrichtian [?largely] derives from the palynomorph
14 record of Seymour Island, Antarctica. The small, spine-bearing dinoflagellate cyst
15 *Impletosphaeridium clavus* is dominant, accounting for up to 99% of the marine
16 palynoflora. Its profusion is interpreted as the accumulation of resting cysts from
17 dinoflagellate blooms associated with winter sea ice decay. Peaks and lows of
18 *Impletosphaeridium clavus* abundance represent particularly cold episodes caused by
19 temporary stratification of polar waters, interposed with warmer periods when the
20 ocean was well-mixed. Immediately prior to the Cretaceous-Paleogene boundary,
21 *Impletosphaeridium clavus* decreased dramatically in abundance, interpreted to
22 represent an early expression of warming associated with Deccan Traps volcanism.
23 Terrestrial conditions inferred from spore/pollen data are consistent with the climate
24 interpretations based on *Impletosphaeridium clavus*, and together provide the highest
25 southern paleolatitude expression of global climate during the Maastrichtian. These

26 **palynomorph data, [?]together with $\delta^{18}\text{O}$ values from macrofossils, support the**
27 **presence of ephemeral ice sheets on Antarctica during the Late Cretaceous, and**
28 **highlight the extreme sensitivity of this region to global climate change.**

29

30 **INTRODUCTION**

31 Following peak warming during the Cretaceous (Turonian) Thermal Maximum (Wilson et
32 al., 2002), global climates cooled during the latest Cretaceous (Friedrich et al., 2012). The
33 well-established, rapid and significant sea level falls at this time requires the [controversial]
34 presence of ephemeral ice sheets on Antarctica (Miller et al., 2005). There is no geological
35 evidence, such as glacial tillites or ice-rafted debris (IRD), of glaciation during the
36 Maastrichtian. Paleobotanical evidence, however, indicates significant cooling in the
37 Antarctic Peninsula region (Francis and Poole, 2002). Here we present new evidence of
38 Maastrichtian climates at 65°S based on palynology, including evidence of particularly cold
39 episodes, using the dinoflagellate cyst *Impletosphaeridium clavus* Wrenn and Hart, 1988
40 emend. Bowman et al. 2013 as a proxy. We present the first high-resolution, quantitative
41 Maastrichtian to Danian palynomorph dataset from Seymour Island, off the northeast tip of
42 the Antarctic Peninsula (Fig. 1). This is the highest southern paleolatitude exposure of
43 sediments of this age in the world.

44

45 **METHODOLOGY**

46 A ~1100 m stratigraphic section comprising homogenous, unconsolidated silty-clays and
47 clayey-silts was measured and sampled at high resolution (i.e. every 0.5 to 2 m) throughout
48 the continuous outcrop of the López de Bertodano Formation (LDBF) on Seymour Island
49 (Figs. 1, 2). Eighty-one sediment samples were quantitatively processed for palynomorphs
50 throughout (Table DR1). The age of the section is Maastrichtian to earliest Danian based on

51 magnetostratigraphy and strontium isotope stratigraphy (McArthur et al., 1998; Tobin et al.,
52 2012). Macrofossil, microfossil and palynomorph evidence together with an iridium anomaly
53 at the Cretaceous-Paleogene (K/Pg) boundary support this age model (Elliot et al., 1994;
54 Olivero, 2011; Bowman et al., 2012; Table DR2). Numerical ages herein are based on
55 Gradstein et al. (2012), which place the K/Pg boundary at 66.0 Ma; this is at 1007.5 m in our
56 section.

57

58 **RESULTS**

59 Dinoflagellate cysts and acritarchs comprise, on average, 70% of the palynofloras below
60 horizon G at 830 m (Table DR1). The small, spine-bearing dinoflagellate cyst
61 *Impletosphaeridium clavus* is dominant, representing up to 99% of the marine palynofloras
62 and attaining up to ~137,000 specimens per gram of sediment. The palynomorph record has
63 been divided by five horizons (A, C, E, G and I; Fig 2) into five intervals (B, D, F, H and J;
64 Fig. 2). Below 830 m, three conspicuous abundance peaks of *Impletosphaeridium clavus* are
65 recorded at horizons A, C and E (143 m, 407 m and 746 m respectively), with intervening
66 intervals of lower concentrations (B, D and F, Fig. 2). The remainder of the marine
67 palynoflora includes typical South Polar Province dinoflagellate cyst genera, such as
68 *Manumiella* spp. (Bowman et al., 2012). Within the same facies, this fluctuating pattern ends
69 abruptly at horizon G with the concentration of *Impletosphaeridium clavus* decreasing to only
70 hundreds of specimens per gram of sediment above 830 m. *Impletosphaeridium clavus*
71 increases again in abundance immediately below the K/Pg boundary, but at relatively low
72 concentrations compared with the lower part of the section.

73 Plant spores and pollen become dominant during interval H (averaging 71% of the entire
74 palynoflora, Table DR1), coincident with the abrupt decrease in concentration of
75 *Impletosphaeridium clavus*. *Laevigatosporites ovatus* (fern) and *Stereisporites*

76 *antiquasporites* (moss) spores, together with *Nothofagidites* spp. (southern beech),
77 *Peninsulapollis gillii* (Proteaceae), *Phyllocladidites mawsonii* and *Podocarpidites* spp.
78 (podocarp conifers) pollen dominate throughout, with rare aquatic fern spores (e.g. *Azolla*
79 spp.) and freshwater algae (e.g. *Botryococcus braunii*). *Nothofagidites* spp. pollen is more
80 abundant and fungal palynomorphs absent at the *Impletosphaeridium clavus* peaks. The
81 distinctive angiosperm pollen species *Tricolpites reticulatus* (Gunneraceae),
82 *Clavamonocolpites polygonalis* (?Chloranthaceae) and *Ericipites scabratus* (Ericaceae) occur
83 rarely between *Impletosphaeridium clavus* peaks A and C, and between peak E and the K/Pg
84 boundary (Figs. 2 and DR1).

85

86 **DISCUSSION**

87 Evidence from neodymium isotopes suggests that deep-water formation, and consequent
88 mixing of the water column, began in the Southern Ocean during the Campanian (Robinson
89 et al., 2010). A relatively mixed water column at this time is supported by a lower planktic to
90 benthic oxygen-isotope gradient during the Maastrichtian compared with the Cenomanian
91 (Huber et al., 2002). Density-driven overturning and wind-driven upwelling, as in the modern
92 Southern Ocean, would have ventilated the entire water column (Mitchell et al., 1991;
93 Sigman et al., 2004). This scenario of a mixed, oxygen and nutrient-rich water column
94 provides the setting for any oceanic change at the Antarctic margin during the Maastrichtian,
95 and may have been the contextual state of the ocean represented by intervals B, D and
96 possibly F in the *Impletosphaeridium clavus* record from Seymour Island (Fig. 2).

97 The acmes of *Impletosphaeridium clavus* at A, C and E (Fig. 2) project from the
98 background levels below 830 m, and are interpreted as reflecting accumulations of resting
99 cysts during periods (lasting perhaps tens of thousands of years) of enhanced seasonal bloom
100 activity (rapid population increases) of the parent dinoflagellates. Dinoflagellate blooms

101 occur today in freshwater and marine coastal environments of the low to mid latitudes in
102 response to enhanced nutrient levels and temperature-driven stratification, primarily during
103 the late spring to early summer (Stipa, 2002; Kudela et al., 2005; Michalak et al., 2013).
104 Dinoflagellate blooms are also known to occur in the modern high northern and southern
105 latitudes associated with the melting of seasonal sea ice, followed by the production of vast
106 quantities of resting cysts (Becquevort et al., 1992; Stoecker et al., 1992). These abundant
107 cysts then fall to the sea floor when the ice melts in late spring or early summer (Harland and
108 Pudsey, 1999).

109 To investigate the striking abundance pattern of *Impletosphaeridium clavus* during the
110 Maastrichtian off the Antarctic margin, we studied geochemical and additional palynofloral
111 data from Seymour Island (Fig. 2). Tobin et al.'s (2012) $\delta^{18}\text{O}$ values from microfossils
112 correlate remarkably well, showing an increase through interval D, then decrease through
113 intervals F and H, mirroring the *Impletosphaeridium clavus* record (Fig. 2). This close
114 correlation of Tobin et al.'s (2012) data with our palynomorph data further strengthens our
115 age model. Overall, heavier $\delta^{18}\text{O}$ values correspond to periods of higher abundance of
116 *Impletosphaeridium clavus* (particularly at horizon E, Fig. 2). We infer from this conformity
117 of pattern that blooms of the dinoflagellate that produced the resting cyst *Impletosphaeridium*
118 *clavus* occurred during periods of particularly cool benthic and intermediate water off the
119 Antarctic Peninsula region during the Maastrichtian.

120 No other marine (or terrestrial) palynomorph exhibits a similar abundance pattern to
121 *Impletosphaeridium clavus* (e.g. Thorn et al., 2009), which suggests that there is a unique
122 environmental factor (not otherwise recorded in the sedimentary record), influencing the
123 success of the dinoflagellate that produced this cyst. We infer from the close correlation of
124 the $\delta^{18}\text{O}$ record that it preferred cool to cold water temperatures, and may even have tolerated
125 near-freezing conditions based on Tobin et al.'s (2012) lowest annually-averaged estimate of

126 ~4-5°C, close to the *Impletosphaeridium clavus* acme at horizon E. In addition, $\delta^{18}\text{O}$
127 measurements from a belemnite (*Dimitobelus seymouriensis* Doyle and Zinsmeister 1988)
128 correlate to our section immediately below the *Impletosphaeridium clavus* acme at horizon C
129 (open arrow “c”, Fig. 2). This suggests a similar intermediate to deep-water mean annual
130 temperature of a cool 6°C at this level with an average annual variability of 5°C (Dutton et
131 al., 2007). These cool intermediate and benthic water temperatures are derived from average
132 annual $\delta^{18}\text{O}$ isotope values, and do not preclude surface water temperatures, especially during
133 the Antarctic winter, dropping below freezing and supporting sea ice development.

134 We infer that the most likely explanation for dinoflagellate blooms in cold, generally well-
135 mixed, shallow marine waters at the Antarctic margin is that they occurred in association
136 with the melting of seasonal sea ice. Although there is no direct modern analogue of
137 *Impletosphaeridium clavus* (Bowman et al., 2013), unsurprising for a cyst from the Late
138 Cretaceous, there are many taxa of small spiny cysts, all comparable in gross morphology,
139 known from sea ice, modern and Quaternary sediment in cold environments of the high
140 southern and northern latitudes (Buck et al., 1992; Stoecker et al., 1992, 1997; Head et al.,
141 2001; Radi et al., 2013). Several of these are characteristic of regions where seasonal sea ice
142 forms (e.g. *Islandinium? cezare* (de Vernal et al. 1989 ex de Vernal in Rochon et al. 1999)
143 Head et al. 2001; Fig. 2) and, comparable to the abundance record of *Impletosphaeridium*
144 *clavus* presented here, some can dominate up to 90% of marine palynomorph assemblages
145 (De Vernal and Rochon, 2011; Radi et al., 2013). The dinoflagellate that produced
146 *Impletosphaeridium clavus* may have had a similar life strategy to that of modern *Biecheleria*
147 *baltica* Moestrup et al. 2009, a brackish water dinoflagellate (also with a small, spiny resting
148 cyst) associated with sea ice cover in the Baltic Sea (Klais et al., 2011; Warns et al., 2012).
149 *Biecheleria baltica* produces vast quantities of resting cysts, which accumulate in benthic

150 cyst beds (at relatively shallow water depths) and promote intense spring blooms by the
151 motile seeding of the euphotic zone (Klais et al., 2011).

152 The reconstruction of pre-Quaternary sea ice from the fossil record in both the Arctic and
153 Antarctic has previously relied on the modern analogue interpretation of *Leiosphaeridia* spp.
154 acritarchs, diatoms, foraminifera and terrigenous IRD (St. John, 2008; Davies et al., 2009;
155 Stickley et al., 2009). Diatoms and foraminifera are sparse in the LDBF (e.g. Harwood,
156 1988). However, the lack of diatoms characteristic of sea ice in the fossil record may be due
157 to a specific paleoecological scenario. For example, early salinity stratification in the modern
158 Baltic Sea favors dinoflagellates over diatoms in the spring blooms associated with sea ice
159 retreat (Klais et al., 2011). No *Leiosphaeridia* spp. or indisputable IRD were recorded from
160 the LDBF. *Leiosphaeridia* spp., which has been used as a proxy for sea ice in the Neogene of
161 Antarctica by Hannah (2006), is present in the Campanian of Humps Island (Wood and
162 Askin, 1992) and may suggest the presence of sea ice in the Antarctic Peninsula region even
163 earlier in the Cretaceous. Habitat difference, e.g. water depth, nutrient availability or levels of
164 oxygenation, may explain their absence in the Maastrichtian record of Seymour Island
165 (Jacobson, 1979).

166 The lack of IRD in our section is perhaps understandable because, based on the presence
167 of diverse and abundant terrestrial palynomorphs (Askin, 1990), the nearest landmass was
168 vegetated and distant from any potential ice caps at high elevations in the Antarctic
169 continental interior. If sea ice was present it may not have been land-fast, preventing the
170 entrainment of out-sized terrigenous IRD. Manoj et al. (2013) noted that IRD is largely
171 absent in the Holocene of the Indian sector of the Southern Ocean, inferring that this is a
172 result (as in past interglacials) of ice sheet accumulation being mainly restricted to high
173 continental interiors. Similarly, and perhaps for the same reasons, no direct evidence of ice

174 from Late Cretaceous sediments was recovered from Ocean Drilling Program Leg 113 cores
175 in the Weddell Sea (Kennett and Barker, 1990).

176 If, as we contend, fossil *Impletosphaeridium clavus* abundances may be used as a proxy
177 for seasonal sea ice, an explanation is required for the episodic, vast accumulations of this
178 cyst that peak at horizons A, C and E (Fig. 2). Dinoflagellates bloom today at the retreating
179 sea ice margin, so variation in the accumulation of fossil resting cysts through geological
180 time in the LDBF must reflect an overprint of an additional longer-term climate or habitat
181 change. One suggestion is these periods of enhanced production of resting cysts may reflect
182 particularly cold climatic phases where the water column became more stratified on the
183 shallow marine shelf, as suggested for deeper waters in the Southern Ocean during the
184 Pliocene (Sigman et al., 2004). These phases, straddling the *Impletosphaeridium clavus*
185 abundance peaks, perhaps lasted tens of thousands of years, and may reflect variations in
186 orbital cyclicality. [agreed – it is a tad speculative – JBR]

187 On the LDBF shelf, which had a slightly fresher surface layer due to riverine inflow, a
188 climatic cooling may have diminished the influence of background temperature-driven
189 mixing, thereby allowing the establishment of a more stable water column in the basin
190 influenced by salinity stratification. During the spring, the formation of an oxygen-rich
191 surface layer, with increased light levels (due to less mixing), and nutrients replenished via
192 buoyant terrestrial inflow and the sea ice melt, would have promoted monospecific blooms of
193 habitat-specific dinoflagellates. Dinoflagellates would still have bloomed during sea ice
194 retreat between these colder spells (i.e. during the middle of intervals B, D and F, Fig. 2), but
195 with less stratification, the light levels would have been lower and the incoming nutrients
196 more diluted throughout the water column, resulting in lower numbers of dinoflagellates and
197 their resting cysts.

198 Additional evidence for Maastrichtian climatic change comes from the terrestrial
199 palynomorph record of Seymour Island, which indicates that the region was forested with
200 southern beech (*Nothofagus* spp.) and conifers (mainly *Podocarpidites* spp.). Most modern
201 relatives of the southern beech grow in wet, cool to cold environments (Read et al., 2010).
202 Below the K/Pg boundary, *Nothofagidites* spp. abundance was lowest during phase H and
203 highest below G, peaking alongside the *Impletosphaeridium clavus* acmes at horizons A, C
204 and E (Fig. 2, Table DR1). By contrast, *Nothofagidites* spp. abundance was lower, although
205 variable, in the intervening warmer phases B and D (Fig. 2). Furthermore, pollen representing
206 specific thermophylic plants such as Gunneraceae (*Tricolpites reticulatus*) appears first
207 during warmer phase B, and much more frequently above horizon E as the region warmed
208 again towards the latest Maastrichtian (Fig. 2, Table DR3). The presence of fungal remains
209 during warm phases B and D, their absence during cold episodes A, C and E, and their
210 increase in frequency of occurrence following E supports evidence for pre-K/Pg warming
211 (Fig. 2). This is consistent with an increase in saprophytic degradation of the terrestrial
212 biomass which grew during the warmer climate. In addition, from the marine macrofossil
213 record, the restricted stratigraphic occurrences of the belemnite *Dimitobelus seymouriensis*
214 Doyle 1988 and the ammonite *Zelandites varuna* Forbes 1846 support our interpretation of
215 short-term warmer intervals (Zinsmeister, 2001; Fig. 2).

216 Previous research supports a generally cool climate in this region during the
217 Maastrichtian. Dicotyledonous wood analysis suggests mean annual terrestrial temperatures
218 of 7°C (Francis and Poole, 2002) and marine molluscan extinction patterns indicate
219 Campanian to Maastrichtian cooling associated with regional shallowing (Crame et al.,
220 1996). Additionally, chemical weathering and sediment maturity suggest probable cold
221 temperate or sub-polar conditions for the Antarctic interior during the Maastrichtian (Dingle
222 and Lavelle, 1998).

223 With no change in facies, the abrupt decrease in abundance of *Impletosphaeridium clavus*
224 across horizon G in the LDBF is interpreted as representing rapid warming that lasted until
225 the K/Pg boundary, which prevented the build-up of winter ice in the northern Antarctic
226 Peninsula region. This coincides with lower abundances of *Nothofagidites* spp. and an
227 increase in the occurrence of thermophilic plants and saprophytic fungal remains through
228 phase H, all suggesting warming. *Impletosphaeridium clavus* increases again in abundance
229 across the K/Pg boundary, but irregularly and in low numbers. Similarly, *Nothofagidites* spp.
230 again increases in abundance into the Danian. This may indicate slight cooling associated
231 with disturbed environments and an erratic recovery period from the K/Pg catastrophe.

232 The palynomorph record from the LDBF of Seymour Island strongly suggests a pattern of
233 at least three particularly cold episodes during the Maastrichtian, followed by an abrupt
234 warming immediately prior to the K/Pg event. This is consistent with other records of
235 Maastrichtian climate trends, which indicate overall cooling until a sudden warming before
236 the K/Pg boundary (Barrera and Savin, 1999; Abramovich and Keller, 2003). From the mid
237 Campanian to ~500 ka prior to the K/Pg boundary, oxygen and carbon isotope excursions
238 indicate two significant cooling episodes (Barrera and Savin, 1999; Friedrich et al., 2012)
239 superimposed on a long-term cooling trend of intermediate waters, particularly in the high
240 latitudes. The first of these, the Campanian-Maastrichtian Boundary Event (CMBE, base of
241 C32N1 to upper C31R), is characterized by a significant global negative carbon isotope
242 excursion, associated with climatic cooling and sea level fall (Friedrich et al., 2009; Jung et
243 al., 2012). Based on magnetostratigraphy, the end of the CMBE correlates with the lower
244 ~500 m of our section, encompassing the cold episodes interpreted at horizons A and C (Fig.
245 2). The second cooling episode recognized from the deep sea occurred from uppermost C31N
246 to the top of C30N (Barrera and Savin, 1999) bracketing the final *Impletosphaeridium clavus*
247 acme during the Maastrichtian (horizon E, Fig. 2). The causes of these oceanic cool phases

248 remain controversial; they have been related to short-term changes in thermohaline
249 circulation and/or the development of ephemeral ice sheets on Antarctica in association with
250 global regressions (Miller et al., 2005; Koch and Friedrich, 2012). Further evidence
251 supporting intervening warmer phases comes from $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ analysis of paleosol
252 carbonates (Nordt et al., 2003; correlated to interval B, Fig. 2) and bulk carbonate (Voigt et
253 al., 2012; interval D, Fig. 2).

254 The earliest records of abundant *Impletosphaeridium clavus* in the James Ross Basin
255 supports the timing of the onset of cold climates in Antarctica during the Campanian
256 evidenced by neodymium isotopes, and the coeval extinction of inoceramid bivalves
257 (Dolding, 1992; Crame and Luther, 1997; Robinson et al., 2010; Bowman et al., 2013). The
258 switch to deep-water production at the Antarctic margin at this time (Robinson et al., 2010)
259 strengthens the link made here between abundant *Impletosphaeridium clavus* and the
260 presence of seasonal sea ice in the Cretaceous because the formation of cold deep waters on
261 the Antarctic shelf today is intrinsically linked to winter sea ice formation (Withworth et al.,
262 1998).

263 Our data interpretations suggest climate may have been cold enough periodically to
264 initiate short-term glaciations of Antarctica, probably restricted to high elevations in the
265 continental interior, during the Maastrichtian. Modeling exploring Cenozoic glaciation
266 corroborates this theory by predicting that significant sea ice forms only in the Southern
267 Ocean *after* the initiation of ice sheets in central Antarctica (DeConto et al., 2007). Hong and
268 Lee (2012) concurred that the Maastrichtian was cold enough to have allowed Antarctic
269 glaciation using estimates of $p\text{CO}_2$ levels below 500 ppm from low latitude pedogenic
270 carbonates.

271 Analysis of our data suggests that a latest Maastrichtian climatic warming (beginning after
272 horizon E, Fig. 2) probably prevented the survival of seasonal sea ice at this paleolatitude,

273 even in the coldest winters. One explanation for this is that at the Antarctic margin, ~2 Ma
274 prior to the K/Pg boundary, the marine and terrestrial fossil record has recorded the beginning
275 of a latest Maastrichtian greenhouse event. Globally, warming through ~500 ka prior to the
276 K/Pg boundary, has been attributed to outgassing associated with the main pulse of Deccan
277 Traps volcanism (Barrera and Savin, 1999; Abramovich and Keller, 2003; Nordt et al., 2003).
278 The earliest (albeit minor) eruptive phase of Deccan Traps volcanism began at the base of
279 C30N (Chenet et al., 2009), coincident with significant changes in the fossil record on
280 Seymour Island. At this level, the final decline in *Impletosphaeridium clavus* numbers began
281 prior to the K/Pg boundary, *Nothofagidites* spp. pollen abundance began to decrease, warmth-
282 loving angiosperm taxa suddenly appeared, saprophytic fungal spores became more prevalent
283 and a decrease in $\delta^{18}\text{O}$ isotope values from macrofossils occurred (Tobin et al., 2012) (Fig.
284 2). We contend that our data from this expanded Maastrichtian section highlight the extreme
285 sensitivity of the Antarctic paleoenvironment to global climatic change.

286

287 **CONCLUSIONS**

288 We provide the first evidence endorsing the presence of seasonal sea ice during the Late
289 Cretaceous at the Antarctic margin. The profusion of *Impletosphaeridium clavus* from the
290 Maastrichtian succession of Seymour Island is considered to represent dinoflagellate blooms
291 and subsequent accumulation of their resting cysts related to winter sea ice decay, much like
292 those produced in abundance today by sea ice dinoflagellates in the Southern Ocean. We
293 consider this as a novel potential proxy for seasonal sea ice and thus Antarctic paleoclimate.
294 Superimposed on a cooling Maastrichtian climate and a generally well-mixed water column,
295 *Impletosphaeridium clavus* acmes suggest particularly cold episodes during temporary
296 stratification of shallow marine waters, interposed with slightly warmer periods when
297 temperature-driven mixing was re-established. Climate interpretations of the terrestrial

298 palynomorph record and $\delta^{18}\text{O}$ isotope analyses from macrofossils support these
299 interpretations. Based on our age model, the palynomorph record herein represents the
300 highest southern paleolatitude expression of global Maastrichtian climate events, including
301 evidence for the end of cooling and perhaps early expression of warming in the latest
302 Cretaceous associated with Deccan Traps volcanism. The Seymour Island palynomorph
303 record of seasonal sea ice for most of the Maastrichtian, in turn, strongly supports year-round
304 glaciation at elevation in the interior of the continent. This agrees with the controversial
305 notion of ephemeral Late Cretaceous ice caps on Antarctica as a causal mechanism for
306 eustatic change and highlights the extreme sensitivity of the high southern paleolatitudes to
307 global climate change.

308

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317

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510 **FIGURE CAPTIONS**

- 511 Figure 1. Location map of Seymour Island, Antarctica, using modern geography.
- 512
513 Figure 2. The abundance of key Maastrichtian palynomorphs from Seymour Island (this
514 study), $\delta^{18}\text{O}$ isotope values and key occurrences of macrofossils (Tobin et al., 2012;
515 Zinsmeister, 2001) and their interpretation regarding regional paleoclimate and possible sea
516 ice presence. Timescale based on the correlation of magnetostratigraphy (Tobin et al., 2012)
517 and strontium isotope stratigraphy (McArthur et al. 1998; Vonhof et al., 2011) to our section
518 using the K/Pg boundary as a datum (updated to Gradstein et al., 2012). Age model verified
519 using biostratigraphical data cited in Table DR2 and an iridium anomaly found at the K/Pg
520 boundary along strike (Elliot et al., 1994). Open arrow “a”, *Pycnodonte* cf. *P. vesiculosa*
521 (Sowerby 1823), 0 m (relative to K/Pg); open arrow “b”, *Linotriconia pygoscelium* [Van – I

522 could not find the author citation for this one – J], bivalve, -200 m; open arrow “c”,
523 *Dimitobelus seymouriensis*, belemnite, -636 m (McArthur et al., 1998). Palynomorph events
524 A to J are discussed in the text. Photomicrograph, i, *Impletosphaeridium clavus*, D5.930.1A,
525 England Finder W65-1; ii, modern *Islandinium? cezare*, holotype, Saint-Césaire, Quebec (SC
526 86-09, slide UQP 199-3B, B15-4; courtesy of Martin J. Head, Brock University). Brown
527 triangles and black dots denote the presence of fungal palynomorphs and pollen comparable
528 with modern angiosperm pollen from plants typical of warm, humid environments (Fig. DR1;
529 Table DR3). The snowflake symbols indicate particularly cold climatic episodes as
530 interpreted from the acmes of *Impletosphaeridium clavus*. *, Chenet et al. (2007, 2009);
531 Robinson et al. (2009).

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535 ¹GSA Data Repository item 20xxxxx, Figure DR1, Tables DR1, DR2 and DR3, and
536 additional references are available online at www.geosociety.org/pubs/ft2009.htm, or on
537 request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140,
538 Boulder, CO 80301, USA.